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Spatial variability of shallow soil moisture and its stable isotope values on a karst hillslope

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ABSTRACT

Soil moisture (θ) and its stable isotope values are two of the most commonly used parameters for studying hydrological processes in soils. Despite their unique ability to aid in distinguishing between soil water evaporation and plant transpiration, the spatial variability of soil water isotope values is not fully understood. In the current study, 10 m × 10 m grids were established within a 90 m × 120 m plot on a highly heterogeneous karst hillslope. Two sampling campaigns were conducted during the early growing season, on April 15, and during the mid-growing season, on August 18, 2011. Stratified soil samples were collected from the shallow soil layer (0–30 cm) to measure θ and its stable isotope values, which were represented by soil water δD values (δD_θ). Related soil properties, land cover and topography were also measured and treated as influencing factors. On both sampling dates, θ decreased with depth, while δD_θ stayed constant for all soil layers and were similar to the most recent rainfall values. Additionally, the variance of δD_θ was smaller than that of θ, especially on August 18 when the most recent rainfalls had similar δD values. The high contrast between θ and δD_θ caused the impacts of other factors on δD_θ to be masked by the impact of the recent rainfall. Soil moisture presented a moderate to strong spatial dependence, which was consistent with the spatial variability of the influencing factors that were significantly correlated with θ. Soil water δD value exhibited weak spatial dependence and random spatial patterns that were different from the other influencing factors. Moreover, significant correlations between these same influencing factors and δD_θ disappeared after a partial analysis with θ as a controlled variable, which means δD_θ was indirectly affected by these influencing factors through θ. This suggests that the spatial variability of δD_θ was being controlled at fine scales. Our results highlight the importance of analyzing the spatial variability of δD_θ, its influencing factors and θ in shallow soil layers separately.

1. Introduction

The O and A horizons are the most active layers in a soil profile, as they are susceptible to rainfall, evaporation and transpiration (Grayson et al., 1997; Brocca et al., 2007). Previous studies reveal the role these horizons play in hydrological processes, such as partitioning of rainfall into surface runoff and infiltration, redistribution of infiltrated rainwater and facilitation of plant transpiration demand (Famiglietti et al., 1998; Chen et al., 2010; Legates et al., 2011). In these studies, soil moisture (θ) and its stable isotope values, were the two most commonly used parameters. Monitoring of θ at different depths can provide information on flow path, infiltration mechanisms and evapotranspiration (Stothoff et al., 1999; Daly and Porporato, 2006; Lange et al., 2010). For some purposes, as when θ monitoring is problematic or for water source identification, a stable isotope technique is the best choice (Nie et al., 2010). However, despite the important role it plays in hydrological research and modeling, there is a lack of knowledge around the spatial variability of soil water isotopes.

Stable water isotopes, which are ubiquitous and occur naturally in water, are ideal for tracing hydrological processes (Ayalon et al., 1998; Lee et al., 2007; Song et al., 2009). Stable water isotopes in soils are affected by the spatial and temporal variations in rainwater (Jayasena et al., 2008; Hu et al., 2013), even though stable water isotopes in rainwater vary slightly and only on small scales (Kato et al., 2013). Transpiration reduces θ but has no effect on soil water isotope values with the main reason for the positive change in isotope values being evaporation (Dawson et al., 2002). Therefore, without the interference of rainfall, a negative correlation between θ and its stable isotope values (especially in shallow soil layer) is expected. In field conditions, soil water storage capacity and plant transpiration vary across space, which make it

Abbreviations: θ, soil moisture; δD_θ, soil water δD values; T, temperature; T_{max}, maximum air temperature; T_{min}, minimum air temperature; Max, maximum; Min, minimum; ET_0, reference evapotranspiration; SD, standard deviation; CV, coefficient of variation; ρ_b, bulk density; CC, clay content; SOC, soil organic carbon; VC, vegetation coverage; EBR, exposed bedrock ratio; RFC, rock fragment content.

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possible for points in space to retain properties of the antecedent rainfall (Jost et al., 2005). Due to unpredictable rainfall recharge and the stable isotope values themselves, correlation between the spatial variability of θ and its stable isotope values should be quite complicated, especially in a highly heterogeneous region.

Geostatistics, first employed in soil science by Campbell (1978), was used to study the spatial variability of sand content and pH. Since then, studies have been conducted to understand the spatial patterns of soil properties and its associated variables (Bruckner et al., 1999; Brocca et al., 2007; Hu et al., 2011). Kriging (Journel and Huijbregts, 1978), a tool in geostatistics, is used to estimate values at unsampled sites, describe the spatial distribution of a variable (Snepvangers et al., 2003; Jost et al., 2005), and reveal correlations to associated factors.

Recently, many studies have been conducted in typically heterogeneous regions (especially in karst regions of southwest China) to reveal the spatial variability of θ and other soil properties (Zhang et al., 2011, 2012; Peng et al., 2013). However, very few, if any, studies have looked at the spatial patterns of soil water isotope values, largely because of the prohibitively large number of samples needed, and the cost associated with analyzing them. Even with reduced costs associated with isotope testing, monitoring of θ is difficult to carry out in shallow, rocky and discontinuously distributed karst soils (Chen et al., 2011; Tokumoto et al., 2014), and it is more reasonable to gather detailed information about the spatial variability of soil water isotope values.

In this study, 10 m × 10 m grids were established within a 90 m × 120 m plot on a typical karst hillslope in southwest China. Stratified soil samples were collected from the shallow soil layer (0–30 cm) to measure θ and its stable isotope values. Related soil properties and environmental factors were also collected. The objectives of this paper were to (1) study the spatial variability of shallow θ and its stable isotope values, and (2) discuss their correlations/differences and influencing environmental factors.

2. Materials and methods

2.1. Site description

The study area is a small catchment (area = 1.14 km^2) located in the Huanjiang Observation and Research Station for Karst Ecosystems under the Chinese Academy of Sciences (24°43′58.9″–24°44′48.8″N, 108°18′56.9″–108°19′58.4″E) in Huanjiang County of northwest Guangxi, southwest China (Fig. 1). A subtropical mountainous monsoon climate dominates with an annual (average value for Huanjiang County from 1986 to 2005) rainfall of 1389 mm and temperature of 18.5 °C (Song et al., 2010). The catchment is a typical karst area with a flat depression surrounded by mountain ranges on three sides with an outlet in the northeast. Elevation in the study area ranges from 272 m to 647 m. Hillslopes are steep with 60% having a grade greater than 25°. Exposed bedrock, underlain by weathered dolomite, covers approximately 30% of the area with thin discontinuous soil containing rock fragments (Chen et al., 2011). The vegetation of this area can be classified into three secondary communities: tussock, shrub, and secondary forest. 70% of the hillslopes are dominated by tussocks and shrub. Secondary forest is only found on the continuous dolomite outcrops or in deep soils at the foot of hillslopes (100 cm in depth). In this study, a typical southeast-facing shrubland hillslope was selected (Fig. 1). Soil depth increased while rock fragment content decreased from upslope to downslope. The vegetation is dominated by shrub with a greater density downslope than upslope.

2.2. Samples collection

Two sampling campaigns were conducted on April 15 and August 18, 2011 to reflect the variations between the early and vigorous growing seasons, respectively. The study area experienced approximately one week of clear days, after the last rainfall, before the two sampling campaigns. The rainfalls occurred on April 8 and August 12, 2011, respectively. Stratified (0–10, 10–20, and 20–30 cm) soil samples were collected by grid (10 m × 10 m) sampling using a 3 cm diameter hand auger within a 90 m × 120 m plot. A total of 130 sampling points were collected on the shrubland hillslope. Each sample was measured for gravimetric θ (expressed by the mass percentage of soil water out of dry soil, %) and water stable isotopes. In total, 390 samples (130 samples at each depth) were collected on each sampling campaign. Samples were placed in capped vials, wrapped in parafilm and stored frozen. During the second sampling period, surface (0–10 cm) undisturbed soil samples were collected by a steel cylinder with a volume of 100 cm^3 and five replicates of disturbed soil samples (130 samples for each) were collected at each point. Exposed bedrock (bare rock without vegetation cover) ratio (EBR) and vegetation coverage (VC), the percentage of vegetation occupying the ground area in vertical projection, were estimated visually within an area of 2 m × 2 m around each sampling point. Elevation and slope were measured by hand-held GPS (Trimble Jono SC, accuracy 2–5 m) and compass readings at each point. Climatic data including maximum (T_{max}) and minimum (T_{min}) air temperature, and the amount of each rainfall event (calculated for a 24-h period between 8:00 am and 8:00 am) were obtained from a standard weather station located in the central depression of the catchment. Stable isotopes were measured from rainwater. During each rainfall event, rainwater was collected in a plastic tank with safeguards to prevent evaporation (Nie et al., 2011) to measure stable isotopes. In this study, eighteen rainfall samples were collected and measured, ten before April 15 and eight before August 18.

2.3. Lab analysis

Soil moisture was measured by an oven-drying method and soil water was extracted by cryogenic vacuuming (Ehleringer et al., 2000; Goebel and Lascano, 2012; Orlowski et al., 2013). The isotopic compositions of soil water and rainwater samples were measured with a liquid water isotope laser spectroscopy instrument (Lis et al., 2008; Penna et al., 2010, 2012; Wassenaar et al., 2014) (Model DLT-100, LGR Inc.). Results are reported in δ notation relative to V-SMOW as

$$\delta D(^{18}O) = (R_{sample}/R_{standard} - 1) \times 1000 \tag{1}$$

where R_{sample} and $R_{standard}$ are the ratio D/H or $^{18}O/^{16}O$ of a measured sample and a standard sample, respectively. The standard deviation for repeat measurements was ±1‰ for δD and ±0.2‰ for $\delta^{18}O$. The δD and $\delta^{18}O$ values of soil water were significantly correlated ($\delta D = 8.43\delta^{18}O + 14.56$, $R^2 = 0.90$) and showed similar trends. Previous studies state that $\delta^{18}O$ values of soil water were greatly affected by soil carbonate but δD values were not (Meißner et al., 2014). Only the results obtained by the analyses of soil water δD values (δD_θ) were introduced in this study.

Undisturbed soil samples were used for bulk density (ρ_b, measured with a gravimetric method) and soil porosity (obtained by $1 - \rho_b / \rho_s$, where ρ_s is the particle density of the soils, which was assumed to be 2.65 g cm^{-3}) measurements. Rock fragment content (RFC), clay content (CC), measured by Matersizer 2000 laser particle size analyzer, and soil organic carbon (SOC), measured with a potassium dichromate heating method (Zhang et al., 2012) were measured using disturbed samples.

2.4. Data analysis

A semi-variogram was used to examine the spatial structure of θ and δD_θ. The semivariance $\gamma(h)$ was calculated with the following formula (Bruckner et al., 1999; Hu et al., 2011):

$$\gamma(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} [Z(x_i) - Z(x_i + h)]^2 \tag{2}$$